Circulation and transport at the south east tip of Greenland

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ABSTRACT

The circulation and related transports at the south east tip of Greenland are determined from direct current observations of a moored current meter array. The measurements cover a time span from June 2004 to June 2006. The net mean total southwestward transport from the mid-shelf (20 km off the coast at 60°N) to the 2070 m isobath (about 100km offshore) was estimated as 17.3 Sv (1 Sv = 1 × 10^6 m³ s⁻¹) with an uncertainty of 1 Sv. The transport variability is characterized by a standard deviation of 3.8 Sv with a peak to peak amplitude up to 30 Sv. The seasonal variability has an amplitude of 1.5 Sv. Frequencies around 0.1 day⁻¹ dominate the signal, although a variability at lower frequency (∼ one month⁻¹) also appears in winter. The coherence between the observed transport variability and the wind stress curl variability over the Irminger Sea is significantly different from 0 at the 95% confidence level for periods greater than 5 days.


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1. Introduction

The circulation southeast of Greenland provides fresh water to the North Atlantic from the Arctic Ocean and therefore plays a key role in the climate evolution. However, South Greenland is not a very accessible region and only few in situ measurements of currents are currently available. Although many hydrographic surveys have been carried out, they rarely provide absolute transport estimates. Hereafter we present the absolute transport of the East Greenland-Irminger Current and its variability over two years deduced from an array of current meters.

![Diagram of North Atlantic subpolar gyre](image)

**Fig. 1.** Schematic circulation of the North Atlantic subpolar gyre inspired from Lavender et al. (2000); Sutherland and Pickart (2008); Holliday et al. (2009); Lherminier et al. (2010); solid lines are for surface currents, dotted lines for intermediate currents, and dashed lines are for bottom currents. The OVIDE section runs eastward from the tip of Greenland to the Rekjanes Ridge, then southeastward towards Portugal (thin black line). The black rectangle shows the moored current meter array.

From the shelf to the central Irminger Sea, the East Greenland Coastal Current (EGCC), the East Greenland Current (EGC), the Irminger Current (IC) and the Deep Western Boundary Current (DWBC) flow south westward, along the shelf and slope. Several descriptions of these currents are given in literature.

The EGCC was re-discovered by Bacon et al. (2002). This 15 km wide and 100 m deep current was centered 10 km away from the East Greenland coast and transports about 1 Sv southwestward from the Denmark Strait to Cape Farewell, carrying both sea ice from the Arctic and continental melting ice from Greenland. Wilkinson and Bacon (2005) emphasized the high variability of its baroclinic transport. Sutherland and Pickart (2008) suggested that the EGCC is a branch of the EGC that forms just south of the Denmark Strait, under the combined influence of the bathymetry and the wind.

The EGC flows from Fram Strait at 80°N to Cape Farewell at 60°N, west of the Greenland Sea, the Denmark Strait and the Irminger Sea. In the Greenland Sea, it transports cold and fresh water from the Arctic Ocean and the Nordic Seas, along with recirculating Atlantic Water (Woodgate et al. 1999). In the Denmark Strait, it tends to mix with the North Iceland Irminger Current and partly feeds the overflow (Rudels
et al. 2002). The lightest part continues along the shelf and is joined by the westward Irminger Current that bifurcates southwards and brings relatively warm and salty water from the North Atlantic Current (Pickart et al. 2005; Sutherland and Pickart 2008).

At the bottom, the DWBC transports the future North Atlantic Deep Water (NADW), composed of the Denmark Strait Overflow Water (DSOW), capped by the Northeast Atlantic Deep Water (NEADW) and the Labrador Sea Water (LSW) (Bacon and Saunders 2010; Sarafanov et al. 2010).

The moorings were placed along a section perpendicular to the bathymetry southeast of Greenland (Figure 1). At those latitudes, several definitions have been proposed to distinguish all the components of the western boundary current. Most of the time (Krauss 1995; Bacon 1997; Lherminier et al. 2007; Holliday et al. 2009), the EGC refers to the whole southward transport off Greenland above the DWBC ($\sigma_0 < 27.8$). However, more detailed studies distinguish the different water masses that contribute to this current; Sutherland and Pickart (2008) used the 34.8 isohaline to separate the EGC from the IC; Wilkinson and Bacon (2005) used the 33.5 isohaline to identify the EGCC, while Sutherland and Pickart (2008) added a velocity criterion for this purpose. In this study, we focus on the transport of the current from the shelf break to the 2070 m isobath, including the EGC and the IC, and will hereafter refer to it as the East Greenland-Irminger Current (EGIC).

No long Eulerian time series are available off southeast Greenland. Only one estimate of the EGIC at 14 Sv was published (Clarke 1984) from a short-term array south of Cape Farewell, reported in the synthetic Table 1 of Pickart et al. (2005). Most of the absolute transport estimates come from hydrographic surveys, where Doppler current measurements were combined with the geostrophic velocities deduced from the density fields. Pickart et al. (2005) find 13.6 Sv above $\sigma_0 = 27.76$ in August 2001; Bacon (1997) find 21 Sv above $\sigma_0 \sim 27.8$ in August 1991, Lherminier et al. (2007), Lherminier et al. (2010) and Gourcuff (2008) find 30.8 Sv in August 1997, 22.4 Sv in June 2002, 29 Sv in June 2004 (above $\sigma_0 \sim 27.8$) and 18 Sv in June 2006 with the same method and limits, emphasizing the high variability of the current. Holliday et al. (2009) find 28.2 Sv above $\sigma_0 \sim 27.8$ (their Table 1) in September 2005. All these estimates refer to the total southward transport off Greenland, from the shelf break to the first current reversal (at about 40$^\circ$W).

Treguier et al. (2006) used numerical simulations to estimate the meridional overturning and transports along the paths of two hydrographic cruises, carried out in 1997 (Fourex) and 2002 (OVIDE) from Greenland to Portugal. They diagnosed the variability of the barotropic transport between the coast of Greenland and 40$^\circ$W along the latitude of the OVIDE section, representative of the EGIC at 60$^\circ$N. blueIn 5-day time series, they found a large high frequency (period less than one month) variability.

The OVIDE (Observatoire de la variabilité interannuelle à décadale en Atlantique Nord) project contributes to the study of the variability of the North Atlantic subpolar gyre circulation based on repeated hydrography and current measurements (Lherminier et al. 2007). In the framework of this project, a moored current meter array was deployed in June 2004 for a period of 2 years, in order to estimate the transport of the EGIC and to put the results of the surveys into the context of this highly variable current. As a western boundary current, the EGIC is indeed an important component of the circulation of the subpolar gyre, that advects the water mixed and densified in the eastern subpolar gyre towards the convection regions in the Labrador Sea (Lavender et al. 2000; Spall and Pickart 2003; Lherminier et al. 2010). These observations are therefore also useful as benchmarks for numerical models (Treguier et al. 2006) and inverse models (Lherminier et al. 2007).

The Ovide current meter array intersects the western boundary currents at the south east tip of Greenland.
(Figure 1). The positions of the moorings were chosen according to the measurements made during Ovide 2002 (Lherminier et al. (2007)). In this study we will confirm that the core of the current is located on the Greenland continental slope, inshore of the 2000 m isobath, where the last mooring stands (Figure 2). The array was planned to be complemented by an array of deep moorings (CFER project) in order to monitor the Deep Western Boundary Current during the same period of time. The deep moorings were finally deployed later, in September 2005 (Bacon and Saunders 2010). Even if the current meter array does not cover the whole area where the current flows southward (from the shelf break to at most 165 km offshore i.e. about 40°W Lherminier et al. (2007)), it covers the most energetic part of the current. The salinity sections in the Irminger Sea (Holliday et al. 2009; Lherminier et al. 2010) show that the warm and salty anomaly carried by the Irminger Current extends between 42.5°W and 41°W, i.e. above the slope down to the 2500 m isobath. Although the array does not intercept all this anomaly, the weak velocities east of 41.5°W (Figure 2) indicate that the transport of this anomaly is well captured by the mooring array.
Fig. 2. **Top:** positions of all the moorings (note that Cb is at the same location as C) and of the hydrographic stations of the Ovide surveys. The black arrows indicate the direction of the along shore ($210^\circ$) and cross shore ($300^\circ$) components. The south tip of Greenland (Cape Farewell) is visible. **Bottom:** mean currents of the 100m-200m layer from the Ship ADCP during OVIDE 2002, 2004 and 2006 surveys (arrows). The arrows are colored in red when they intersect the current that is actually measured by the mooring array, and in white when they are outside. The transport section is represented by the black thick line.
2. Data and methods

a. Data

Five current meter moorings (A to E) were set out in June 2004 over the continental slope off Greenland inshore the 2000 m isobath and immediately south of 60°N in order to monitor the transport of the currents (Figure 2, 3 and Table 1). The mooring array was recovered in June 2006. For servicing purposes, moorings C and E were recovered in September 2005 and immediately redeployed at the same location for mooring C, renamed Cb after redeployment, and closer to the shore for mooring E renamed Eb (Figure 3). In addition, a mooring was deployed by the National Oceanography Center at Southampton (NOCS), referred to as A0 on Figure 2 and presented (as mooring “A”) in Bacon and Saunders (2010). Since A0 was deployed 15 months after the other moorings, it will not be used in the transport computations.

![Diagram of moorings and depth](image)

**Fig. 3.** Cross-section of the moorings showing the depth of the instruments. The grid used to calculate the transport is drawn in light gray: cell separations are determined by the mid-distance between current meter positions (thin black lines). For the ADCPs at moorings C(Cb) and E(Eb) the grid vertical limits are determined according to the cell thicknesses (16 m and 4 m for C(Cb) and E(Eb) respectively) and only one vertical limit out of two is plotted. In order to closely follow the bathymetry, each cell was decomposed horizontally using a resolution of 1 km or less.

Moorings A to D were equipped with 15 Aanderaa RCM8 current meters, complemented by an upward looking 75 kHz RD Instrument ADCP at the top of moorings C and Cb (Figure 3). Moorings E and Eb were frames lying on the sea floor and equipped with upward looking 150 kHz RDI ADCP (Figure 3). All moorings were equipped with SeaBird Inc. SBE16+ (Seacat) or SBE17 (Microcat) mounted at about 200 m depth.
Table 1. Moorings used in the present analysis; ID is the mooring name and measuring temperature, salinity and pressure. Mooring B Seacat was lost during recovery. Mooring A0 was equipped similarly, with 2 current meters and a Seacat at the top (Table 1), and 3 current meters deeper, which are presented in Bacon and Saunders (2010).

Full calibration details of the RCM8 Aanderaa current meters are given in Branellec and Lherminier (2009). Current speeds are thought to be accurate to ±0.01 m s\(^{-1}\) and current direction to ±10°. Sampling rate was set to 1 hour. Data return was good (about 90%). Some speed measurements were missed due to stalled rotors (see Table 1). At the top of mooring B, the current meter velocity slowed down in the last five months of the time series. The comparison with other current meters, especially with the one below, showed that this trend was most likely due to some biofouling on the rotor. It was corrected, assuming a linear trend on the modulus, by conserving the measured angle and the mean vertical gradient of the velocity (Branellec and Lherminier (2009)).

The 75 kHz ADCP at moorings C and Cb provided current measurements between 520 m and 40 m depth.
(nominally) with a 16 m resolution, a 30 minute sampling rate and an accuracy better than 0.02 m s$^{-1}$. Data is missing at the end of the deployment of mooring C (see Table 1), due to a memory failure (Summer 2005). During the missing period (Table 1), the velocities of this ADCP were set to their two-year average in transport computations. At moorings E and Eb, the 300 kHz ADCP provided current measurements between 190 m and 20 m depth with a 4 m resolution, a 30-minute sampling rate and an accuracy better than 0.01 m s$^{-1}$. The ranges of the ADCPs were reduced between January and April due to lack of reflectors in the surface layers during this period.

Owing to particularly energetic events, the mooring line inclination varied significantly over time. The instantaneous pressure of each instrument was thus computed based on the Seacat and Microcat accurate pressure measurements and mooring geometry (Siedler and Paul 1991). Finally, gaps in current records were filled following the method of Beckers and Rixen (2003) and Alvera-Azcarate et al. (2005).

All data has then been low passed with a 2-day$^{-1}$ cut-off frequency and subsampled daily to eliminate tides and other high frequency variability not of interest in this analysis. This method does not affect the low frequency variance content.

Hydrographic, Lowered ADCP (LADCP) and Ship ADCP (SADCP) surveys were conducted during the Ovide 2004 deployment and Ovide 2006 recovery cruises on N.O. Thalassa and R. V. M. S. Merian respectively (see Billant et al. (2006) and Lherminier et al. (2003) for details of the CTD and LADCP data processing and Figure 2 for station locations). Temperature and salinity accuracy were estimated to be better than 0.001°C and 0.003. LADCP data from the downward looking 150 kHz and upward looking 300 kHz RDI ADCP, both mounted on the rosette were processed following Visbeck (2002). For further reference, the temperature and density sections at deployment and recovery are presented in Figure 4, and the LADCP sections in Figure 5. SADCP data, processed with the Cascade software (Gourcuff et al. 2006), are only used in Figure 2 to illustrate the average shape of the current near the surface.

b. Methods

Previous observations show that the western boundary current system is strongly guided by the bathymetry at all depths. This is why the mooring section was oriented perpendicular to the isobaths, in the direction $120^\circ$ to $300^\circ$. The components of the velocities will be presented in the cross-shore direction ($300^\circ$, $u_{300}$) and along-shore direction ($210^\circ$, $u_{210}$), as shown in Figures 2 and 7. Note that the along-shore direction coincides with the mean current direction and the direction of the first principle axis of variance. The transport across the mooring section is calculated using $u_{210}$ and is positive south-westward. The distances along the mooring section are all computed from the Greenland coast at 60.1°N; 40.15°W, knowing that data collected during the Ovide cruises are located inside an area where there is a good correspondence between the bathymetry and the distance from this point.

In addition to averages and variances, statistics of the data presented in Table 2 show the integral time scales for each level of measurements for the along-shore component, following the method of Teague et al. (2005): the auto-correlation function of the detrended time series are integrated to the first zero-crossing. The statistical error $\varepsilon$ of the mean estimate is the ratio $\sigma/\sqrt{N_{df}}$ of the standard deviation ($\sigma$) over the square-root of the number of degrees of freedom ($N_{df}$), the latter being the total duration of the experiment divided by the integral time scale. The Eddy Kinetic Energy (EKE) per unit of mass is classically calculated as $EKE = \frac{1}{2}(u^2 + v^2)$.
with \( u' = u - \bar{u} \) and \( v' = v - \bar{v} \), the bars representing the time average. Since the duration of the time series (above 9 months) and the 2-day filtering are compatible with the method of Dickson (1983), all the statistics of Table 2 can be used as a complement to their charts. To calculate the transports, the section covered by the moorings was decomposed using a grid, which cell separations are determined by the mid-distance between current meter positions. Each current measurement is then multiplied by the surface of the associated cell. Note that the shallowest cells extend up to the surface. In order to get a realistic area of the bottom cells, each cell was decomposed horizontally using a resolution of 1 km or less (depending on the distance between the moorings). The grid is plotted on Figure 3. It covers the whole water column from 59.931°N; 42.803°W to 59.604°N; 41.628°W, i.e. from 20 km to 98 km from the Greenland coast; the section crosses the isobaths 190 m to 2070 m. During the second year of deployment, mooring Eb was placed closer to Greenland, most likely in the EGCC. However, as the minimum of current between the EGCC and the EGC (Sutherland and Pickart 2008) could not be captured correctly by the mooring array (see Figure 2), the velocities cannot be

<table>
<thead>
<tr>
<th>ID</th>
<th>Lat (°N)</th>
<th>Lon (°W)</th>
<th>Bottom (m)</th>
<th>&lt;D&gt;</th>
<th>Nominal Depth (m)</th>
<th>( \langle u_0 \rangle ) (m s(^{-1}))</th>
<th>( \langle u_0 \rangle ) (m s(^{-1}))</th>
<th>( \langle T \rangle ) (°C)</th>
<th>( \langle T \rangle ) (°C)</th>
<th>( &lt;\text{T pot.} ) (m)</th>
<th>( &lt;\text{T pot.} ) (m)</th>
<th>EKE (cm(^2))</th>
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<td>2.24</td>
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* ADCP cell
realistically interpolated between mooring Eb and mooring D, the closest one. This is why data of mooring Eb was replaced here by the mean velocity measured by mooring E the previous year. The final transport discussed here is estimated across the section that is prescribed by the position of moorings A to E, where the linear interpolation is validated by the SADCP datasets.

The bathymetry was deduced from the swath of a multibeam echosounder (Simrad EM120 of the M.S. Merian) in the vicinity of the moorings A0 to D, up to 59.83°N. Further North, we used the Smith and Sandwell bathymetry (Smith and Sandwell 1997).

Although all calculations are carried out using the grid defined earlier, the vertical sections shown in Figures 8 and 11 are plotted using the Data-Interpolating Variational Analysis (DIVA) software (Troupin et al. 2008) that is well suited to complex frontiers (here, the bathymetry) and helped to better illustrate the regional dynamics on a more regular grid in space and time ($dx = 2$ km, $dz = 50$ m and $dt = 1$ day to 1 month). Mean data of moorings Eb and A0 constrain the frontiers of the DIVA interpolation.

3. Structure and variability of the measured currents

![Figure 4](image)

**Fig. 4.** Potential temperature measured in June 2004 and June 2006 are shown on the left and right panels respectively, along the sections plotted at the bottom-left (software from Schlitzer (2008)). The vertical black lines correspond to Ovide cruise stations. The main isotherms (thin black lines) and isopycnals (thin white lines) are superposed. The 34.8 isohaline is the red thick line. Note that sections are slightly different from the mooring section; in 2004, the section was not really perpendicular to the bathymetry, leading to a slope that looks smoother on this projection. The mooring positions (black dots) are represented according to their respective bottom depth since the circulation is strongly guided by the bathymetry.

Snapshots of the hydrographic properties (Figure 4) and of the currents as measured by LADCP (Figure 5) are provided by Ovide surveys in June 2004 and 2006, limited westward by the ice-edge. Following Sutherland et al. (2009); Myers et al. (2009), the 34.8 psu isohaline divides the polar water carried by the EGC and the EGCC from the warmer and saltier water brought by the Irminger Current. On those late-spring snapshots, mooring E is the only instrument capturing the polar water and, according to the surface intensified velocities ($> 40$ cm s$^{-1}$, Figure 5), this mooring appears to measure the EGCC in June 2004. However, the position of the EGCC is highly variable and possibly linked to the ice-edge (Gourcuff 2008). Similarly, the frontier between the EGC and the Irminger Current at 34.8 is very variable in space and time (Sutherland et al. 2009).
The current section (Figure 5) shows another velocity maximum near the surface that is present on both sections and is associated with the EGIC. It extends from the surface to 1300 m depth and is centered on the 1600-1700 m isobaths. Although these sections are not exactly the same, the guidance of the current by the bathymetry is confirmed by the data. The mooring positions were therefore located according to their respective bottom depth on the different sections. As already noticed by the subsurface currents on Figure 2, the core of the current is in the vicinity of mooring B. However, its intensity varies, with velocities in June 2006, stronger by about 10 cm s\(^{-1}\). The current sections confirm that the cross-shore velocities are much weaker than the along-shore component in the EGIC, showing that the array is reasonably perpendicular to the main stream at all depths.

The upper tail of the DWBC appears as a relative maximum of velocity at the bottom (also shown in Bacon and Saunders (2010)), present in June 2004 but not in June 2006. When defined below the isopycnal \(\sigma_\theta = 27.8\) (Figure 4), close to 3°C isotherm, the DWBC encompasses the three deepest current meters of the array (on moorings A and B). Considering the analysis of Bacon and Saunders (2010), it is unlikely that the mooring array sampled the DSOW, the Irminger Sea deepest water. So, the measured portion of the DWBC is mainly composed of NEADW.

Our objective here is to compute the transport of the Western Boundary Current and report on its variabil-
ity, without distinguishing the water masses that would need more salinity data and a finer spatial sampling. Part of the EGCC is probably included in our estimates, and the portion attributed to the DWBC will be quantified.

Velocities are stronger at mooring B (C) 63% (22%) of the time, confirming that most of the time, the core of the EGIC is centered near mooring B as observed in June 2004 and 2006. Time series of the current, measured at five different depths on mooring B, are presented in Figure 6. The high-frequency variability is visible at all depths. This variability is particularly strong in winter, with even a few short-term current reversals, which affect the whole water column.

![Fig. 6. Time series of currents at mooring B; first column: time series (black) of along-shore component $u_{210}$ (positive southwestward); cross-shore component $u_{300}$ (gray, positive northwestward); The seasonal cycle estimated by least-squares fitting an annual period cosine to the along-shore velocity is also plotted (dash line); second column: current rose plots, with daily averages in grey, mean vector in red and 300° axis in black; third column: means and standard deviations of the speed as a function of depth.]

Rose plots, figuring the daily and average currents, show the narrow dispersion around the mean direction: at all depths, the flow is strongly aligned with the direction of the local topography. The along-shore component coincides with the first principal axis and accounts for 80% of the variability.

As was anticipated from the snapshots in Figure 5, the EGIC is intensified towards the surface, with mean currents over 30 cm s$^{-1}$ above 600 m depth and at about 10 cm s$^{-1}$ below 1400 m.

The seasonal variability is surface-intensified (Figure 6). Its amplitude reaches 6 cm s$^{-1}$ at 206 m where it explains 13% of the variance. It decreases downward to 3 cm s$^{-1}$ at 614 m and 1 cm s$^{-1}$ at the bottom.
where it explains only 1% of the variance. The phase of the signal is such that the amplitude of the seasonal signal reaches a maximum (minimum) in January (July) and follows that of the wind stress curl annual cycle in the Irminger Sea. Those results are qualitatively in accordance with those reported by Fischer et al. (2004) from the analysis of a moored array in the Western Boundary Current at the exit of the Labrador Sea. Fischer found a maximum transport in November-December and showed that the seasonal cycle is confined to the upper Labrador Current, close to the shore. In our observations the seasonal cycle seems also trapped on the slope, its amplitude being 2 cm s\(^{-1}\) or less at mooring A. At 206 m, a seesaw shape in winter 2005 (a slowing down in January-March followed by a sudden recovery in April, Figure 6) reminds us a similar signal observed in the Western Boundary Current at the exit of the Labrador Sea (Fischer et al. 2004). This seasonal and intra-seasonal variability, also visible in the transport estimates, will be discussed further in the transport section.

No significant trend was observed over the two years of observation.

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**FIG. 7.** Map of the record-length mean current vectors. Instrument depths are color coded.

The spatial structure of the boundary current, averaged over the two years, is illustrated on a horizontal map (Figure 7) and a vertical section of the along-shore component (Figure 8a), while the total EKE section (Figure 8b) shows its variability. The corresponding values are reported in Table 2. The statistical error associated with the along-shore mean current, between 0.2 et 2 cm s\(^{-1}\) (Table 2) stands below 6%. As previously depicted for mooring B, the time-averaged currents are all directed towards the south-west and are closely aligned with the bathymetry from surface to bottom, particularly at moorings C and D that are located on the steep continental slope, and at mooring E on the shelf. The mean currents are relatively intense with maximum
velocities up to 35 cm s\(^{-1}\) at the shallowest levels of mooring B and C. At the bottom, a relative minimum of velocity below 13 cm s\(^{-1}\) is observed between the 1000 m and 1800 m isobaths. Offshore, at mooring A, the mean velocity is still intense, with amplitudes of 20 cm s\(^{-1}\) from surface to bottom. The current there is nearly barotropic. The mean current on the shelf is weaker with velocities of about 10 cm s\(^{-1}\).

Close to the ocean floor the velocity core appears to shift eastward. This feature is related to the DWBC, defined below the 3\(^\circ\)C isotherm, and was observed in the LADCP section of June 2004 (Figure 5). It appears that while the velocity minimum at mooring B is found at the bottom, the relative minimum of velocity at mooring A is found at intermediate depths, although not very contrasted (still close to 20 cm s\(^{-1}\)), confirming the difficulty of choosing a reference level in this area. Bacon and Saunders (2010) (their Figure 5) have shown that the bottom velocity decreases east of mooring A, before increasing again eastward. The lower velocities, observed at the eastern limit of the array section (constrained by A0 mooring average), is thus consistent with this statement (Figure 8a).

Similarly to the mean current, the EKE (Figure 8b, Table 2) is also intensified in the upper 600 m, particularly near the EGIC core. It decreases rapidly to low values below 1000 m. The position of the EKE maximum, closer to the shelf break than the EGIC core, may be the consequence of the higher variability of the current western edge, but our interpretation of this pattern is very limited by the lack of information on the shallow part of mooring B. However, it certainly decreases offshore, towards and beyond mooring A (mooring A0 was used for the plot and its EKE values are reported in Table 2).

A spectral analysis of the along-shore velocities \(u_{210}\) at mooring B (Figure 9) shows that the energy at low frequency decreases with depths. It confirms the already mentioned vertical decay of the seasonal signal (Figure 6), and is consistent with the decay of the integral time scale (Table 2) from 7.8 days at 206 m to 1.2 day near the bottom. A peak of energy is observed at about 10 days at all depths. At the other moorings, the spectra are similar to those of the deep levels of mooring B, even in the top 700 m where the integral time
Fig. 9. Spectra of along-shore component $u_{210}$ at Mooring B; spectra are calculated for overlapping pieces of 128 days duration; 90% Confidence interval is indicated; the legend relies colors to depths and corresponding integral time scales.

scales are lower than at mooring B. By contrast, the high-frequency variance, with periods less than 5 days, is very similar at all depths for all moorings (see mooring B on Figure 9).

4. Transports

a. Total transport of the Western Boundary Current

Using the along-shore velocities and the grid described above (Figure 3), we calculate a 2-year time series of the transports across the section defined by the array (top to bottom). This daily time series is presented in Figure 10, along with the result of a (30 day)$^{-1}$ cut-off frequency low-pass filter. The mean transport of the Western Boundary Current is 17.3 Sv, with a standard deviation of 3.8 Sv. Since the integral timescale is found at 4.8 days, the mean transport stands between 16.7 and 17.9 Sv, using a 95% confidence interval. Note that the correction made on the velocity time series at the top of mooring B increased this average by only 0.1 Sv.

The time series of the transport of the Western Boundary Current shows an energetic high frequency variability, dominated by the signal in the East Greenland-Irminger Current. However, the transport never reversed, meaning that the current is always oriented towards the southwest. No significant trend can be detected over the 2 years of measurements.

As explained by Bacon and Saunders (2010), the very western edge of the DWBC is included in our transport
estimates. Using the 3°C as the upper limit of this current above (Figure 8), it accounts for 2-2.3 Sv in the mean transport (Figure 10), with no apparent trend. The correlation between the transport in the DWBC and the transport above the DWBC is 0.3; this weak correlation is however significantly different from zero ($p_{\text{value}} = 10^{-15}$).

The surface-intensified seasonal variability, identified in the along-shore velocity, translates into a seasonal modulation of the transport which maximum (minimum) is reached in winter (summer) (Figure 10). We estimate $1.5 \pm 0.4$ Sv for its amplitude, with a 95% confidence interval.

The spatial structure of the extreme values in transport are illustrated in Figure 11 (right panel). The minimum transport of 3 Sv was observed on 10 June 2005, with a quite narrow southward core and a strong return flow offshore from the surface down to 1600 m. The maximum at 33 Sv was observed on 21 March 2005, with a wide current centered above the steep slope showing velocities of 0.8 m s$^{-1}$ in the sub-surface core. Several similar, though less extreme, patterns exist in the transport time-series. They are transient, less than 5-day long (Figure 10), and not necessarily associated with extreme transports in the bottom layer. Their spatial structures suggest that they may be due to advection of surface-intensified mesoscale structures (Figure 11).
The spatial patterns of the variability are investigated through a decomposition in Empirical Orthogonal Functions (EOFs) computed as the eigenfunctions of the normalized covariance matrix of the alongshore current variability (Figure 12). The first three EOFs that account for 40%, 18% and 12% of the eddy kinetic energy capture the main patterns of the variability (Figure 12). EOF 1 describes the variability above the continental slope, at the western edge of the EGIC. This variability appears uncorrelated with that at mooring A and to a lesser extend B mainly described by EOFs 2 and 3, respectively. The maximum (minimum) transport of 33 Sv (3 Sv) observed on 21 March 2005 (10 June 2005) corresponds to a moment when the first (second) principal component (PC) is at an extreme and the second (first) principal component at a relative minimum. The spectral analysis of the PC time series shows that most of the low frequency energy ($< 10^{-1}$ days$^{-1}$) is captured by PC 1 (Figure 12). In the following section, we show that PC 1 is significantly coherent at the 95% confidence level with the wind stress curl over the western Irminger Sea.
Fig. 12. First 3 EOFs of the current meter array time series and spectral analysis of the associated principal component in variance preserving form. The percentage of total variance explained by each EOF is indicated in the bathymetry. Contours represent the EOF amplitude. The grey number is the percentage of variance explained by the EOF at a given depth.

b. Spectral analysis

A variance preserving form spectrum of the whole time series (Figure 13) shows that the energy is concentrated between 8 and 16 days. While the variance of the signal corresponds to \((3.8 \text{ Sv})^2\), the variance integrated in the 2-30 day band is responsible for \((3.2 \text{ Sv})^2\), i.e. 71% of the total variance. In Figure 14-b, a time-frequency analysis of the total transport is carried out using a wavelet spectral analysis (Torrence and Compo 1998) \(^1\). It confirms that the variability is the most energetic within the periods from 8 to 15 day. Overall, most of the transport variance is found at periods of less than a month.

The transport time series seems to be affected by a stronger intra-seasonal variability in winter. Once accumulated for periods above 8 days in the wavelet analysis, the variance (Figure 14-c) shows two significative

\(^1\)Wavelet software was provided by C. Torrence and G. Compo, and is available at URL: http://paos.colorado.edu/research/wavelets/software.html
maxima in early January 2005 and late January 2006. These variability maxima occur in periods that coincide with the maxima of the seasonal signal. In winter variability is twice as energetic as in summer.

c. Estimation of errors

There are different sources of errors in the estimations of transport by the mooring array. First, the ADCP mounted on mooring C showed that the current is occasionally surface intensified (Figure 11, right panels), and this pattern cannot fully be captured by mooring B, since the top instrument lies below 200 m depth. This would lead to a slight under-estimation of the current, about 0.1 Sv if extrapolated from the shear seen by the upward-looking ADCP of the nearby mooring C from 200 m depth to the surface.

Another source of uncertainty is due to the spatial sampling of the moored array. It is particularly relevant in the vicinity of the deep moorings A and B, where some narrow velocity anomalies could have been missed. Based on SADCP sections made with a 2 km resolution, an anomaly of 0.1 m s$^{-1}$ integrated over 15 km and down to 1800 m is plausible, and leads to an error of 2.7 Sv in the instantaneous transport estimate from our moored array. This is probably the main source of error in each value of transport, but because of its random nature, it is not expected to affect the average over 2 years.

Finally, using the 2004-2005 average of the transport at mooring E (0.51 Sv) for the 2005-2006 period was found to be insignificant when compared to the EGIC variability. On the shelf, only a very small portion of
the EGCC was included in our estimates.

Taking those possible sources of error into account, a standard deviation of 4.7 Sv and a possible underestimation of 0.1 Sv were calculated, so that the final standard error associated with the 17.3 Sv mean transport between 59.931°N; 42.803°W and 59.604°N; 41.628°W is 1 Sv when using an integral time scale of 4.8 days and a 98% confidence interval.

5. Atmospheric forcing

One expects that some of the time variability of the transport of the EGIC is due to the wind stress. When comparing the transport of the WBC in two simulations of a high resolution model, one with climatological forcing fields, and one with daily varying realistic forcing (from European Center for Medium range Weather
Treguier et al. (2006) have shown that the variability at periods of less than one month is a response to wind and heat fluxes. The atmospheric variability is indeed particularly strong here due to the presence of barrier winds and tip jets (Moore and Renfrew 2005). To investigate a possible relationship between the wind forcing and the EGIC transport variability, we have first computed the coherence and phase in the frequency domain between the wind stress variability from ECMWF ERA 40 at the location of the moored array and the EGIC transport variability for two period bands (5-25 days and 25-128 days). The coherence and phase are the module and the phase of the cross-spectrum. Since no significant coherence was found between the local wind stress and the EGIC transport variability, we looked in a second step for larger scale coherence patterns and considered the wind stress curl variability in the Irminger Basin (Figure 15). We find significant coherences at the 95% confidence level between the wind stress curl and the EGIC transport variability in both periods bands. At lower frequency (25-128 day periods), the area of maximum coherence covers a large part of the western Irminger Sea while at higher frequency (5-25 day period) it is more localized and the coherence is weaker. The phase of the coherence (not shown) is 1-2 days for the 5-25 day periods and 4-5 days for the longer periods, suggesting a rapid response of the western boundary current to Ekman pumping variability. The first EOF of the current meter time series captures most of the coherent response to the wind stress curl variability for the 25-128 day period band, showing that the latter affects mainly the western edge of the EGIC (not shown). The coherence is weak at higher frequency.

**Fig. 15.** Coherence of the wind stress curl with the 2004-2006 EGIC transport time series for the 5-25 day period band (left panel) and 25-128 day period band (right panel). Regions where the coherence is not significantly different from 0 at the 95% significance level, or those that are occasionally sea ice covered, are in white.

### 6. Discussion

The Ovide mooring array, located east of the south tip of Greenland at 59.75°N, measured the current from June 2004 to June 2006. The array was conceived to estimate the transport of the southward East Greenland-Irminger Current, but it included also the western edge of the Western Boundary Current below (under 3°C), and part of the East Greenland Coastal Current on the shelf (13% and 3% of the mean transport measured by the array respectively). Since mooring Eb, near the core of the EGCC, covered only the last 9 months of the
period, when no data was available to document the minimum of current between Eb et D (mooring E was not in place any more), Eb data were not taken into account in this analysis. A more specific study will be done in the future to document EGCC transport.

The section defined by the array extends from 20 km to 98 km when counted from the Greenland shore (60.1°N; 43.15°W), perpendicularly to the slope, covering isobaths 190 m to 2070 m. The current, strongly guided by the bathymetry, is found to be reasonably orthogonal to this section at all time.

Although there is some additional southward flow east of the array (Figure 2 and Figure 16), the core of the EGIC is captured mostly by mooring B, sometimes by mooring C, with a sharp decrease eastward: currents at mooring A are about 10 cm s$^{-1}$ less than those at mooring B (Table 2). Note also that temperatures averaged over the first 900 m decreases by 0.5°C from B to A, showing that A stands on the eastern edge of the warm and salty anomaly.

In the core of the EGIC, currents intensify toward the surface with mean velocities of 37 cm s$^{-1}$ at 300 m depth and 20 cm s$^{-1}$ at 1000 m (mooring B, Figure 8). The eastern edge of the EGIC lies over the western edge of the DWBC, leading to a barotropic structure of the current profile at mooring A.

The EGIC is one of the boundary currents of the subpolar gyre, continued by the West Greenland Current (WGC) and the Labrador Current downstream (Figure 1), as attested by drifting buoys and subsurface floats (Lavender et al. 2000; Fratantoni 2001; Cuny et al. 2002; Jakobsen et al. 2003; Reverdin et al. 2003). Those currents are characterized at the surface by a relatively weak EKE compared to the energy of the mean current (MKE), which contrasts with the larger EKE/MKE ratios observed in the interior of the Labrador and Irminger gyres (Jakobsen et al. 2003, their Figure 11b). Following Dickson (1983), we have estimated the EKE/MKE ratio in the 0-800 m and 800-1800 m layers, excluding the outer shelf mooring E. We find $\langle EKE \rangle = 86 \text{ cm}^2 \text{s}^{-2}$ and $\langle MKE \rangle = 324 \text{ cm}^2 \text{s}^{-2}$ in the 0-800 m layer, leading to a ratio of 0.27. In the 800-1800 m layer we find a ratio of $37/116 = 0.32$. As pointed out by Dickson (1983), weak ratios of EKE over MKE “tend to be found at sites occupied by intense and/or constrained flows.”

Downstream, however, the boundary current is destabilized by topographic irregularities, as observed in the WGC offshore Cape Desolation (Fratantoni 2001; Eden and Böning 2002; Bracco et al. 2008; Chanut et al. 2008), resulting in higher values of EKE/MKE. Based on a study of the energy balance in high resolution numerical simulations this increase in EKE is explained by Eden and Böning (2002) as a result of the barotropic instability of the West Greenland Current related to the convergence of the geostrophic streamlines upstream of Cape Desolation. Further downstream, at the exit of the Labrador Sea, Fischer et al. (2004) find again small ratios of EKE over MKE.

The total transport associated to the Ovide mooring array is found at $17.3\pm1 \text{ Sv}$ with a standard deviation of 3.8 Sv. No trend emerged over the 2 years of measurements. The seasonal signal amounts to 3 Sv peak to peak, explaining 8% of the transport variance. The intra-seasonal variability dominates at all depths: 71% of the transport variance is explained by the variability at periods between 2 and 30 days. The variability is strengthened in winter, particularly in 2005 (Figures 10, 14), although the transport averaged over winter months (Dec-March) is the same in 2005 and 2006: 18.4 Sv. Both the classical and the wavelet spectral analysis show two significant peaks of variability at 3 and 10-11 days in the total transport.

Many similarities could be found between our observations and those reported by Fischer et al. (2004) at the exit of the Labrador Sea. They found 1) a peak of current variability at periods of 10-20 days 2) a
There are different sources of variability in the transport measured by the current meter array. First of all, the study of Treguier et al. (2006), as well as the reported coherence between the EGIC transport variability and the wind stress curl over the Irminger Basin, suggest that the wind is a major forcing of the observed variability.

In addition, part of the variability is likely to be associated with anomalies created upstream. These anomalies can result from the recirculation cells observed upstream (Bacon et al. 2008), from the variability of the 2 Sv Spill Jet described by Pickart et al. (2005) or from the deep cyclonic eddies created by the Denmark Strait Overflow (Käse et al. 2003), since the array captures about 13% of the DWBC transport.

And finally, transport estimates at Cape Farewell by Holliday et al. (2007) showed that at Eirik Ridge (south of the mooring array), about one third of the EGIC was retroflexed back into the central Irminger Basin. This estimation was based on the total southward current off Greenland. If one supposes that this ratio is steady, and that the recirculation affects mostly the eastern edge of the current, then about 20% of the transport measured by the current meter might be affected by this recirculation. Whether this recirculation is the source of the variability seen in the transport measured by the current meters is impossible to determine at this moment, since no synoptic array was available in the West Greenland Current or further east in the Irminger Sea.

Additional numerical model studies are nevertheless needed to better understand the mechanisms of the observed variability and to explain the apparent similarities between our observations and those at the exit of the Labrador Sea reported by Fischer et al. (2004).
APPENDIX

Comparison with geostrophic estimates of the transport

![Graph showing comparison of vertically integrated cumulative transport from the Greenland shelf to southward transports.](image)

**Fig. 16.** Comparison of vertically integrated cumulative transport from the Greenland shelf (59.93°N; 42.8°W) to 59.4°N; 41°W; distance is counted from 60.1°N; 43.15°W, so that there is a good correspondence between the bathymetry of the Ovide and the mooring sections. Heavy solid lines are from hydrographic inversions: June 2002 is in green, June 2004 in blue, June 2006 in red; moorings estimates in June 2004 and 2006 (heavy dashed lines) are plotted with their 95% confidence interval (light dashed lines). Unlike previous figures, but in agreement with Lherminier et al. (2007), the negative values are for southward transports.

Once accumulated from the shelf offshore, the transport calculated from the mooring array is comparable to absolute geostrophic transports estimated by inversion during the Ovide surveys of 2002 (Lherminier et al. 2007), 2004 (Lherminier et al. 2010) and 2006 (Gourcuff 2008). The transport derived from hydrographic inversions are representative of averages over the duration of the survey (about 20 days). Since the paths are not exactly the same (Figure 2), the comparisons are performed as a function of distance consistent with the topography, as done before. In Figure 16, the first and last values of the mooring cumulative transports (after filtering, Figure 10) are plotted, along with the absolute transports of the surveys. The survey values are higher, both in 2004 and 2006 by about 2 Sv, but this is not incompatible with the standard error calculated in the inverse model (2 Sv) and estimated in the 30-day mooring average (1 Sv). This difference might be related to the different locations of the hydrographic and mooring sections, but can also be interpreted as a sampling artefact. The important result is that we find a good consistency between these independent estimates when
we observe that in June 2006, the integrated transport was 4-5 Sv weaker than in June 2004 in both cases.
The results reported by Holliday et al. (2009) over 5-7 September, 2005 shows about 21 Sv flowing south when accumulated from the shore to 100 km offshore (i.e. the eastern limit of our grid) in their Figure 5. Despite the uncertainties, it seems that the value is compatible but again above the values deduced from the current meter array (Figure 10).
The decrease of 4-5 Sv seen in the total transport between June 2004 and June 2006, when observed on the time series in Figure 10, is associated with an intra-seasonal variability, but not with a trend.

Acknowledgments.

Nathalie Daniault is supported by the ”Université Européenne de Bretagne”, Pascale Lherminier by Ifremer and Herlé Mercier by the CNRS. We acknowledge Ifremer and INSU who funded this project. Thank you to Olivier Peden, André Billant, Stéphane Leizour and Olivier Ménage who prepared and recovered the moorings, and to Pierre Branellec and Michel Hamon who processed the data. We are grateful to Captain Piton and the crew of N/O Thalassa (Ovide 2004), and to Captain Von Staar and the crew of M.S. Merian (Ovide 2006), and to all the scientists that helped in the mooring deployment and recovery. We are particularly grateful to Sheldon Bacon and his staff, who took time to renew moorings C and E in September 2005 (cruise D298 on the RRS Discovery), and to recover mooring Eb in September 2006 (cruise D301 on the RRS Discovery). Thank you to Sheldon Bacon and Nicolas Ducousso for their constructive remarks, and to Peter Saunders for his careful analysis of the deep current meter of mooring A and his helpful suggestions about this manuscript.
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